Seasonal variations of hydrological cycle components in the Mississippi River basin from a MAPS version with snow and frozen soil physics

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Abstract

A coupled atmospheric/land-surface model covering the conterminous United States with an associated 1-hour atmospheric data assimilation cycle, the Mesoscale Analysis and Prediction System (MAPS), has been improved to include snow and frozen soil physics. The new aspects of the land-surface model are described in this paper, along with detailed one-dimensional (1-D) tests. These tests show that the MAPS 1-D soil/vegetation/snow model is capable of providing accurate simulations over multi-year periods at locations with significant snow and frozen soil processes. The performance of the full 3-D model/assimilation system running at a 40-km resolution over a 9-month period from November 1997 through July 1998 is then examined. Soil moisture and temperature at multiple levels have been cycled in MAPS since April 1996 and snow water equivalent depth and snow temperature since March 1997. Cycling of these fields gives estimates that are physically consistent with the evolution of atmospheric fields over this fairly long period and are vastly improved over climatological estimates. Precipitation and surface temperature fields show good agreement with monthly analyses from the National Climatic Data Center (NCDC) in the 9-month comparison. Soil moisture fields and hydrological cycle components such as precipitation-minus-evaporation, snow accumulation, snow melt, and surface runoff are also examined from the MAPS cycle and are qualitatively reasonable and mutually consistent. Some needed improvements for MAPS are indicated by these experiments, including reducing moderate spin-up in 0-1 hour precipitation forecasts by improving initial cloud/moisture fields and eliminating biases in convective precipitation over warm oceans and nearby land areas. Overall, the results indicate a system with continuous cycling of soil and snow fields and frequent data assimilation, accurate model physics, and good overall model performance, can provide good seasonal as well as short-range estimates of unobserved components of the hydrological cycle.

1. Introduction

The development of improved capabilities for climate prediction and climate impact assessment for various factors requires a better understanding of the time and space variability of water and energy budgets over continental and subcontinental regions. To meet this goal, it is necessary to develop and validate high-resolution coupled atmospheric/land-surface models and also to develop methods for initializing them. The initialization consists of assimilation of diverse observations in the atmosphere and surface, consistent with the complex physical relationships in these systems. This issue was discussed in the Scientific Plan for the GEWEX (Global Energy and Water Cycle Experiment) Continental-scale International Project (GCIP) [*World Meteorological Organization* 1992, *International GEWEX Project Office* 1993].

The Mesoscale Analysis and Prediction System (MAPS) [*Benjamin et al.*, 1997, 1998] is a state-of-the-art coupled model and data assimilation system operating over the conterminous United States (US) and producing grids for the GCIP. MAPS was developed at the NOAA Forecast Systems Laboratory (FSL) where it is run on a real-time, continuous basis. It also has been implemented in a fully operational mode at the National Centers for Environmental Prediction (NCEP) as the Rapid Update Cycle or RUC. The 40-km, 40-level MAPS has been producing Model Output Reduced Data Set (MORDS) grids for GCIP since May 1996. MAPS is unique in that it provides these grids from an ongoing assimilation cycle, including evolution of soil moisture and temperature. This cycling of soil fields has been ongoing since April 1996, so that the MAPS cycle is, in essence, providing seasonal records of these mostly unobserved fields. In May 1997, the analysis interval was shortened from 3 hours to 1 hour, meaning that hourly data such as profiler and surface observations are now being assimilated in their full temporal frequency. From that time onward, the analyzed state of each hourly MAPS forecast consists of the previous 1-hour forecast (first guess) of all fields; atmospheric, multi-level soil, and clouds, are corrected by observations valid in a 1-hour window near the analysis valid time. These observations used in MAPS include those from rawinsondes, surface atmospheric observation stations, commercial aircraft, wind profilers, and geostationary satellites. A summary of the characteristics of the 40-km MAPS is provided in more detail by *Benjamin et al*. [1998, 1997].

MAPS uses an isentropic-sigma hybrid vertical coordinate, which is advantageous for moisture transport and resolution of temperature, moisture, winds, and other atmospheric variables in the vicinity of fronts. It also has high vertical resolution near the surface regardless of terrain elevation. A 6-level soil/vegetation/snow scheme has been incorporated into MAPS to improve its predictions of surface fluxes and atmospheric boundary-layer properties by explicitly predicting soil moisture and temperature in a data assimilation cycle rather than depending on climatological soil moisture values, which can be seriously in error during and after dry or rainy periods.

The MAPS model produces forecasts of both grid-scale and (parameterized) convective precipitation. These forecasts (0-3 hours or 0-1 hour) are the moisture input to the soil model. The grid-scale precipitation can fall to the ground as either solid (snow or graupel) or liquid (rain) phase [*Brown et al.* 1998]. Convective precipitation is assumed to be entirely rain. The liquid phase is infiltrated into the soil at a rate that cannot exceed the maximum infiltration rate, with the excess going into surface runoff. The solid phase is accumulated on the ground/snow surface and is unavailable for the soil until the melting process begins.

The soil/vegetation/snow model was first tested in long-term integrations in a Project for the Intercomparison of Land-surface Prediction Schemes (PILPS) [*Schlosser et al*., 1998) mode using a 1-D configuration of MAPS/RUC. After it was implemented into the full MAPS assimilation scheme, monitoring of the hydrological cycle started. MAPS-produced monthly accumulated precipitation, snow accumulation, snow melt and surface runoff for the whole domain, and Mississippi River basin in particular, were produced for the period of November 1997 through July 1998. MAPS precipitation and temperature are compared with those observed for this period and overall behavior of the MAPS hydrological cycle over the GCIP continental area is discussed.

The main question addressed in this paper is whether a coupled atmospheric/land-surface model, constrained by hourly assimilation of atmospheric observations to follow the evolution of the atmosphere accurately, can provide a realistic evolution of hydrological fields and time-varying soil fields that are not observed over large areas. A prerequisite for success is that the soil/vegetation/snow component of the coupled model, which is constrained only by atmospheric boundary conditions and definition of fields such as vegetation type and fraction and soil type, must be sufficiently robust to avoid drift over long periods of time. One-dimensional tests of land-surface process models initialized with and verified against multi-year soil data sets with observed atmospheric forcing provide a controlled environment to examine model behavior. Therefore, as part of our investigation, considerable attention has been given to such 1-D tests.

In section 2 of this paper, the most recent version of the soil/vegetation/snow model in the coupled atmospheric/surface MAPS forecast model, including frozen soil processes, is described. Detailed tests of this land surface process model in a 1-D framework are presented in section 3. Based on the full coupled MAPS model and its associated hourly data assimilation cycle, seasonal variations of hydrological cycle components in the Mississippi River basin have been calculated and analyzed, as shown in section 4. Concluding remarks are presented in section 5.

2. Soil/vegetation/snow model description

The MAPS/RUC soil model contains heat and moisture transfer equations together with the energy and moisture budget equations for the ground surface, and uses an implicit scheme for the computation of the surface fluxes [*Smirnova et al.* 1997 a,b]. The heat and moisture budgets are applied to a thin layer spanning the ground surface and including both the soil and the atmosphere with corresponding heat capacities and densities (Fig. 1). A concept for treating the evapotranspiration process, developed by *Pan and Mahrt* [1987], is implemented in the MAPS/RUC soil/vegetation scheme.

Soil temperature and volumetric water content, as predicted by the soil model, have been incorporated into the MAPS/RUC assimilation cycle. Because a high-frequency, national domain precipitation analysis is not yet available in real time, it is necessary to depend on MAPS/RUC precipitation forecasts for precipitation input. Lack of observed precipitation data and soil moisture information in real time implies that the predicted soil fields, particularly deep soil moisture, are vulnerable to “model drift” due both to inadequate precipitation input and to deficiencies in the soil model itself or in the soil properties that it uses. These potential problems are addressed in sections 4 and 5.

2.1 Parameterization of snow accumulation and snow melting processes

To improve MAPS/RUC prediction of skin temperature and surface air temperature in the cold season, and to avoid significant errors which may result even at short time scales from inaccurate forecasts of snow cover, a snow physics parameterization and snow cycling component has been added to the soil/vegetation scheme initially described in *Smirnova et al*. [1997b]. The snow physics package accounts for the processes of snow accumulation on the ground surface and snow melting.

When snow is present, snow is considered to be an additional upper layer of soil that interacts with the atmosphere, significantly affecting the surface characteristics. The properties of snow are quite different from those of soil. High values of albedo reduce the amount of absorbed solar radiation, and the small thermal diffusivity in snow reduces coupling with temperatures in the soil layers below. As a result, the skin temperature may be much cooler where there is snow cover. Further, the atmospheric stratification frequently becomes stable with inversions near the ground.

The snow model contains a heat-transfer equation within the snow layer together with the energy and moisture budget equations on the surface of the snow pack. The integrated form of heat budget equation on the snow can be written as follows

$$(\rho_a c_p \Delta z_a + \rho_{sn} c_{sn} \Delta z_{sn}) \frac{\partial T_{sn}}{\partial t} = \{R_n + H_{rain} - H - F - L_s[E_{dir}(1-\sigma_f) + E_c\sigma_f + E_t\sigma_f]\}\Big|_{\Delta z_a} - G_{sn}\Big|_{-\Delta z_{sn}} \tag{1}$$

where T_{sn} is temperature of the snow surface, R_n is net radiation flux, H_{rain} is heat brought in to the ground surface by the liquid phase of precipitation, H is sensible heat flux, F is latent heat required to melt snow, L_s is latent heat of sublimation, E_{dir} is the evaporation rate over the bare soil, E_c is the evaporation rate from the canopy, E_t is transpiration, G_{sn} is the heat flux into the snow, and σ_f is fraction of grid box covered by vegetation. (In the cold season σ_f can be very small or even zero.) A detailed list of symbols is given in the appendix.

This budget equation is applied to the layer from the middle of the first layer in the atmosphere to the middle of the snow layer. If snow cover is deeper than a threshold value (currently set equal to 7.5 cm), then the energy budget is still applied to the top half of the threshold layer. Applying the equation in this manner is supported by the known fact that the largest thermal gradient below the snow surface is near the top of the snow layer due to the small value of thermal diffusion in snow. It means that for deep snow cover, the snow layer below the threshold value is essentially isothermal. Thus, the heat flux into the snow in (1) is defined by

$$G_{sn} = \nu_{sn}\frac{T_{sn} - T_s}{h_{sn}}, \qquad (2)$$

where ν_{sn} is thermal conductivity of snow (set equal to a constant value of 0.35 $WK^{-1}m^{-1}$), T_{sn} is the "skin" temperature, T_s is the temperature at the soil-snow interface, or snow temperature at the threshold depth, h_{sn} is the snow cover depth, or if the snow depth is higher that the threshold value, h_{sn} is set equal to the threshold value.

Direct evaporation from the snow surface is the most significant among the three evaporation components in (1), because even if some leaves remain on trees, and the vegetation fraction is non-zero, the evapotranspiration is suppressed by the cold temperatures. Snow evaporates at a potential rate unless all of the snow layer would evaporate before the end of the time step. In this case the evaporation rate is reduced to that which would just evaporate all the existing snow during the current time step. Melting at the top of the snow layer occurs if the energy budget produces temperatures higher than the freezing temperature (0^{o} C). In this case, the snow temperature is set equal to the freezing point, and the residual from the energy budget is used to melt snow (F>0 in (1)). Water from melting snow infiltrates into the soil, and if the infiltration rate exceeds the maximum possible value for the given soil type, then the excess water becomes surface runoff.

The accumulation of snow on the ground surface is provided by the microphysics algorithm of the MAPS/RUC forecast scheme [*Reisner et al.*, 1998, *Brown et al*., 1998]. It predicts the total amount of precipitation and also the distribution of precipitation between solid and liquid phases. The subgrid-scale ("convective") parameterization scheme also contributes to the liquid precipitation. With or without snow cover, the liquid phase is infiltrated into the soil at a rate not exceeding maximum infiltration rate, and the excess goes into surface runoff. The solid phase in the form of snow or graupel is accumulated on the ground/snow surface and is unavailable for the soil until melting begins. The integrated finite-difference form of the water budget on the interface between snow and soil is written as

$$\rho_l \Delta z_s \frac{\partial \eta_g}{\partial t} = -W_s\big|_{-\Delta z_s} + \{(1-\sigma_f)I_m + \sigma_f D - E(1-\sigma_f) - E_t\sigma_f\}\Big|_{\Delta z_a} \quad , \tag{3}$$

where η_g is the volumetric soil moisture content at the soil-snow interface, W_s is the soil moisture flux through the middle of the top soil layer, I_m is the infiltration flux into soil originated from snow melt and liquid portion of total precipitation flux, E is the flux of total moisture content in the atmosphere, D is the excess water dripping from the vegetation canopy onto the soil when the canopy is saturated, and it is defined by

$$D = \begin{cases} P_l - E_c, & C^* \ge S' \\ 0, & C^* < S' \end{cases} . \tag{4}$$

Here, P_l is the flux of liquid precipitation, S' is the saturation water content for a canopy surface, and C^* is the actual canopy water content.

2.2 Parameterization of processes in frozen soil

Frozen soil plays a significant role in the hydrology of many regions, decreasing infiltration into the soil and causing large runoff rates from otherwise mild rainfall or snowmelt events. Significant runoff over saturated and unprotected soils may cause extreme erosion that may threaten agricultural productivity and construction projects. The need to control runoff and erosion and to determine sensitivity of these processes to soil properties and types of crops and vegetation covering the ground surface has generated much attention to modeling of freezing and thawing processes among hydrologists and soil scientists. Many methods to predict the depth and permeability of frozen soil dependent on the interrelated processes of heat and moisture transfer within the soil have been developed [*Harlan*,1973; *Fuchs et al*.,1978; *Jame and Norum*, 1980; *Flerchinger and Saxton*, 1989]. Many of the models have a high degree of sophistication and accuracy in predicting soil freezing depths and profiles of temperature, water and ice in the soil. However, simultaneous heat and mass transport requires an iterative procedure for numerical solution, making these models computationally expensive and not practical at the present time for coupling with atmospheric models used operationally for weather forecasting. For these coupled operational forecast models, a parameterization of frozen soil physics is needed that describes freezing and thawing processes is needed and is simple enough to be computationally efficient. In winter 1997-98, such a parameterization of frozen soil physics was incorporated into the MAPS coupled atmospheric/surface forecast model and assimilation cycle after extensive testing in a 1-D framework. This 1-D testing is described in section 4, and the behavior in the coupled 3-D MAPS is examined in section 5.

Lukianov and Golovko [1957] proposed two simplifying assumptions for frozen soil physics: that the only significant phase change occurs between liquid water and ice, and that there is no flow of ice. Based on this assumption, a 1-D heat balance equation for a soil layer in which both latent and sensible heat are transported can be written as [*Harlan*, 1973]

$$C\frac{\partial T}{\partial t} - L_f S_{li} = \frac{\partial}{\partial z}\left(\nu_f \frac{\partial T}{\partial z}\right), \tag{5}$$

where L_f is the heat of fusion, ν_f is the thermal conductivity of the potentially frozen soil, and S_{li} is the rate of liquid mass transformation into ice defined as

$$S_{li} = -\rho_l \frac{\partial \eta_l}{\partial t}. \tag{6}$$

where ρ_l is the density of water, and η_l is the volumetric content of liquid phase in soil. Applying the definition of liquid mass transformation rate to (5), the heat balance equation becomes

$$C_a\frac{\partial T}{\partial t} = \frac{\partial}{\partial z}\left(\nu_f\frac{\partial T}{\partial z}\right) \tag{7}$$

where C_a is called the apparent heat capacity and is equal to

$$C_a = C + \rho_l L_f\frac{\partial\eta_l}{\partial T}. \tag{8}$$

The slope of the soil freezing characteristic curve, $\partial\eta_l/\partial T$, can be obtained from *Flerchinger and Saxton* [1989] under the simplifying assumption of zero solute concentration in the soil solution:

$$\eta_l = \eta_s\left[\frac{L_f(T-273.15)}{gT\Psi_s}\right]^{-1/b}, \tag{9}$$

where η_s is the volumetric moisture content at saturation, Ψ_s is the moisture potential for saturated soil. The heat capacity of the soil is calculated according to the weighted contribution of the dry soil, liquid water and ice:

$$C = (1-\eta_s)C_s + \eta_l C_l + \eta_i C_i. \tag{10}$$

The thermal conductivity ν_f for soils with partially frozen water is defined from [*Pressman*, 1994]:

$$\nu_f = \nu\left(1 + \frac{\rho_i}{\rho_l}\eta_i\right), \tag{11}$$

where thermal conductivity for unfrozen soils ν is calculated as it is described in *Smirnova et al.* [1997b].

The water balance of a soil layer at subfreezing temperatures can be written in terms of total water mass concentration as

$$\frac{\partial\Theta}{\partial t} = \frac{\partial}{\partial z} D_f \frac{\partial\Theta}{\partial z} + \rho_l \frac{\partial K_f}{\partial z} \quad , \tag{12}$$

where

$$\Theta = \begin{cases} \rho_l \eta_l, & (\eta_i \neq 0) \\ \rho_l \eta_l + \rho_i \eta_i, & (\eta_i > 0) \end{cases} \tag{13}$$

is the total water mass content, D_f is diffusional conductivity in the frozen soil, and K_f is hydraulic conductivity in the frozen soil. According to experimental data [*Jame and Norum,* 1980], the presence of ice in soil disrupts the established flow paths and therefore reduces the water flow speed, and the impending factor is assumed to be a function of total ice content. This experimental data showed that this factor may increase exponentially from 1 for ice-free conditions to 1000 when ice content is greater that 20%. Results from *Bloomsburg and Wang* [1969] showed that hydraulic conductivity is zero if $(\eta_s - \rho_i / \rho_l \eta_i) < 0.13$. The formulations of hydraulic and diffusional conductivity used in MAPS are written as follows [*Pressman,* 1994]:

$$K_f = K_s \left(\frac{W - \rho_i \eta_i}{W_s - \rho_i \eta_i} \right)^{2b+3} \left(1 - \frac{\rho_i \eta_i}{W_s - W_r} \right)^a \tag{14}$$

$$D_f = K_s \frac{\partial \Psi_f}{\partial \eta} \tag{15}$$

$$\Psi_f = \Psi_s \left(\frac{W_s - \rho_i \eta_i}{W - \rho_i \eta_i} \right)^b \left(\frac{W_s}{W_s - \rho_i \eta_i} \right)^c \tag{16}$$

Here, K_f and D_f are hydraulic and diffusional conductivities in the frozen soil, respectively, K_s is the hydraulic conductivity at saturation, Θ_s and Θ_r are densities of maximum possible and minimum values of soil moisture content, respectively, b is the exponent in the *Clapp and Hornberger* [1978] parameterization, and a and c are empirical parameters. All these parameters are the function of the soil type, except for *a* and *c*, which are set equal to 1 and 3, respectively, for all soil types. In case there is no frozen soil water, (14)-(16) transform into formulations used in MAPS previously and described in *Smirnova et al.* [1997b].

3. One-dimensional experiments

The MAPS soil/vegetation/snow model was tested off-line in a 1-D setting before incorporation into the MAPS/RUC three-dimensional (3-D) forecast model. Its snow physics package and frozen soil physics parameterization, in particular, needed data from a site with a significant winter season, including data about snow cover on the ground surface. As part of the increasing interdisciplinary effort fostered in part by GEWEX, such data sets have been made available for sites in Russia [*Vinnikov and Yeserkepova*, 1991; *Robock et al.*, 1995; *Schlosser et al.*, 1997]. Experiments with the 1-D version of the MAPS soil/vegetation/snow model at these sites are described below.

3.1 Experiments for Valdai, Russia

The data set most suitable for this 1-D testing was from an observation site at Valdai, which is located in a climatic zone of Russia with significant seasonal variations and persistent snow cover from November until April. This data set includes continuous atmospheric forcing data for 18 years. The Valdai data set has been used for the most recent phase of the ongoing, internationally based Project for Intercomparison of Land surface Parameterization Schemes (PILPS) phase 2d which focused attention on processes of the cold season, which are considered to be of great importance for global climate simulations [*Schlosser et al.*, 1998]. The MAPS 1-D model participated in the PILPS phase 2d intercomparison, along with many other 1-D land surface process models. The MAPS results are described in detail below.

In the Valdai experiment, the model simulates moisture and heat transfers inside the soil, and interaction processes between the ground/snow surface and the atmosphere, including surface fluxes, snow accumulation, and snow melting, as driven by atmospheric forcing. The data sets have a 3-hour frequency, and are interpolated to 30-minute intervals (the model time step) as prescribed by PILPS. The first year of simulations was repeated until an equilibrium state was reached, i.e., when the result is no longer dependent on the initial conditions. The simulated soil moisture, surface runoff, evapotranspiration, and snow water equivalent are verified against the observed data to evaluate the performance of the snow-melting and frozen soil physics algorithms.

The MAPS results of the Valdai experiment show reasonable performance of the snow physics package, and also demonstrate significant impact of frozen soil physics on the hydrological regime of the Valdai catchment during the cold season, and, in particular, in the spring and fall when thawing and freezing of soil moisture occurs. They also indicate the sensitivity of snow physics package to the incoming longwave radiation, which is addressed below briefly and described in details in Schlosser et al. [1998].

a. Effect of frozen soil physics on soil parameter profiles for a specific year.

The presence of ice in soil affects soil properties such as thermal conductivity (Eq. (8), (10) and (11)), and hydraulic and diffusional conductivity (Eq. (14), (15) and (16)). The results from the Valdai experiments using MAPS versions with and without frozen soil physics illustrate the typical effects of ice phase change in soil in this climate regime. By the middle of spring, the soil often has a complicated structure with several melted and frozen layers. Figure 2 depicts the temperature profile obtained from MAPS with frozen soil physics for April 15, 1981. It indicates that the beginning of spring 1981 was warm enough to melt ice in the soil down to more than 1 m deep, although there is still a frozen layer below. After the warm period, cold weather returned, a common occurrence in the Valdai region, and the top 10 cm of soil were frozen again. The temperature profile without consideration of freezing and melting processes in soil is quite different (Fig. 2, dot-dashed line). The changes in soil temperatures happen faster in this experiment due to the fact that latent heat of fusion has been ignored. Temperatures have warmed to above freezing in the entire soil domain (3 m thick) during the warm period. With the return of below freezing temperatures in the atmospheric forcing, they have decreased below freezing in the top layer, but the frozen layer is twice as deep compared to the experiment with soil moisture freezing. There is a sign of warming up in the very top layer in the version of MAPS without frozen soil physics, while the other version does not show this.

The behavior of apparent heat capacity (Fig. 3a) and thermal diffusivity (Fig. 3b) is consistent with the temperature profile shown in Fig. 1 for the frozen soil physics experiment. The apparent heat capacity is temperature-independent at above the freezing point, and increases abruptly by several orders of magnitude when ice formation begins. The increase of heat capacity is larger if the temperature is closer to the freezing point, and after all available water in the soil is frozen it returns to values defined by Eq. (10), because the rate of liquid mass transformation S_{li} becomes zero. Thermal diffusivity, defined as the ratio of thermal conductivity to apparent heat capacity, decreases noticeably when freezing occurs (Fig. 3b), indicating that the propagation of a temperature wave in the soil during freezing or thawing is much slower than in unfrozen soil.

Comparison of soil moisture profiles obtained from the two versions of MAPS for the middle of each season (Fig. 4a-d), show considerable difference for the top 1-m layer in the middle of winter, less difference in the middle of spring, and practically no difference in the middle of summer and fall. In the cold period when part or all available moisture in soil is frozen (and with the assumption that there is no flow of ice), the soil moisture profiles are not changing much in the frozen layers, keeping total moisture content high in the top 1.5 m and low in the bottom layer, as is typical for late fall in the Valdai region. In the MAPS no-ice version, soil moisture is transported when temperatures are below the freezing point at the same rate as in the warm seasons, distributing soil water more uniformly over the soil domain and making the top 1.25 m significantly drier. During the melting season, the soil receives less melted water in the experiment with frozen soil physics if there is ice in the top two layers. Ice presence reduces the maximum infiltration rate or even terminates infiltration when the difference between the saturation value of volumetric soil moisture and ice content becomes less than 0.15. As a result, most melted water goes into surface runoff in the MAPS model with parameterization of frozen soil physics, and the soil moisture profiles from the two versions of MAPS in the middle of spring (Fig. 4b) become closer than in the middle of winter (Fig. 4a), and for the rest of the warm season the difference between them stays negligible.

Diffusional and hydraulic conductivity, which are the functions of total soil moisture content and ice content (Eq. (14)-(16)), show the largest differences between the two MAPS versions in the bottom frozen layer. In this layer, higher ice concentration plays a more important role in the calculation of hydraulic and diffusional conductivities than does soil moisture content (Fig. 5 a,b). The movement of liquid water in this area is significantly slower in the experiment with parameterization of frozen soil physics. The unfrozen layer and the thin top frozen layer with low ice content have closer values of the diffusional (Fig. 5a) and hydraulic (Fig. 5b) conductivities, which are dependent primarily on the soil moisture profiles.

b. Effect of frozen soil physics on long-term averages of soil properties

The performance of the frozen soil physics parameterization in MAPS is also studied from a climatological viewpoint for the 18-year PILPS 2d period. In these experiments, the 1-D model was run for the 18 year period with atmospheric forcing after the equilibrium initialization for the first year. The annual cycles over this period are averaged for different variables in Fig. 6.

The skin temperature shows no significant difference between versions of MAPS with and without frozen soil physics, although in winter, the 18-year averaged daily skin temperatures are slightly cooler with consideration of freezing processes inside soil (Fig. 6a) due to smaller thermal conductivity of soil with ice compared to soil with water. Similar thermal regimes on the ground surface define practically the same amounts of snow accumulation in the two versions of MAPS (Fig. 6b). The snow physics algorithm demonstrated reasonably good simulation of the average date when snow accumulation begins and snow melting ends. The depth of snow cover averaged over 18 years is slightly underestimated (Fig. 6b) and, in another experiment, was found to be sensitive to the amount of incoming longwave radiation (20% lower values of incoming longwave radiation in winter allow the MAPS model to produce more accurate values of snow accumulation, runoff, and the date when all snow is melted [*Schlosser et al.*, 1998]). Monthly accumulated total evaporation is closer to the lower end in the range of observed values, except for the spring time; overall, the discrepancies between the two MAPS versions are not significant (Fig. 6c).

The components of the hydrological cycle are more affected by consideration of processes in the frozen soil (Figs. 6d, 6e). The soil moisture in the top 1-m layer averaged over the 18-year period (Fig. 6d) verifies fairly well against observations, demonstrating moist conditions from October until April and drying out in summer. Soil moisture simulation in winter is less accurate when hydraulic properties in the frozen soil are not changed compared to the warm season. The spring maximum associated with the snow melting process is reflected in both versions of MAPS, but it is more realistic with frozen soil physics because of the reduced capability to infiltrate melted water into the still frozen soil. Due to a surface runoff increase, the total runoff (Fig. 6e) is higher with frozen soil physics when melting of snow occurs. The surface runoff in mid-winter is quite small, and the underestimated storage of soil water in the top meter of soil (Fig. 6d) is explained by the higher drainage of soil water through the lower boundary without frozen soil physics. This higher drainage is reflected in the wintertime higher amounts of total runoff from the 1-m thick layer without frozen soil physics (Fig. 6e). However, when temperatures rise above the freezing point in late spring, this deficiency disappears fairly quickly due to the overestimated amount of infiltrated water from snow melt, and the two models both perform with sufficiently good accuracy (Fig. 6d).

3.2 Results from six mid latitude stations with varying climatic regimes

The data from six stations located in the different climatic regions of the former Soviet Union, provided by Adam Schlosser (pers. comm.) and described by *Vinnikov and Yeserkepova* [1991] and *Robock et al.* [1995], are also excellent for testing of frozen soil physics parameterization. Five of these stations are located in Russia (Khabarovsk, Kostroma, Tulun, Ogurtsovo, and Yershov) and one (Uralsk) is in Kazakhstan. The procedure for experiments conducted for these sites is the same as in the Valdai experiment described above; the atmospheric forcing is available for only a 6-year period (1978-1983).

Figures 7a-7f depict observed and simulated soil moisture in the top one meter of soil over the 6-year period for these six stations. The two versions of the MAPS model (with and without frozen soil physics) both capture the main features in the seasonal variation of soil moisture and also demonstrate consistency with precipitation events and periods of active snow melting. The moisture conditions vary from humid at Khabarovsk and Kostroma to semiarid at Uralsk and Yershov. In the Khabarovsk experiments (Fig. 7a), the agreement of both experiments with observations is generally good, and the model is able to capture the significant drying out of soil in summer. For Kostroma (Fig. 7b), the values of soil moisture storage in the top one meter are generally underestimated, although the seasonal variations are also simulated fairly well. In dry (Fig. 7c - Uralsk, Fig. 7d - Yershov) and moderate (Fig. 7e - Tulun, 7f - Ogurtsovo) climatic conditions, the models also demonstrate good performance, the version with frozen soil physics being generally slightly more accurate in the spring thawing period.

In addition to verification of soil moisture evolution, the data from these six Russian stations give a unique opportunity to verify thermal processes within the soil by comparing model freezing depth against observations, a crucial test for a frozen soil physics parameterization. The variety of soil thermal conditions among these stations makes such verification especially valuable. The model freezing depth is determined with low accuracy due to low vertical model resolution in the deeper layers of soil. It is estimated by linear interpolation of temperature between the levels and finding the depth where it becomes equal to the freezing point. The deepest level at which temperature turns from below freezing to above freezing is considered to be the freezing depth in the model. However, even this crude estimate of the freezing depth is informative in regard to the performance of the frozen soil physics algorithm.

The typical difference in freezing depth comparisons for all six stations (Fig. 8a-f) is the reduction of the freezing depth when the model includes soil water freezing. The release of energy from the freezing process slows the cooling of soil layers until the moment when there is no more available water to freeze. As a result, the slopes of the freezing depth curves are less steep at the beginning of the cold season in comparison with the version of MAPS without frozen soil physics. And in some cases the freezing depth curves follow very closely the observations and reflect the oscillations evident in the observations. However, the value of the freezing depth is not always accurate, being often overestimated in humid Khabarovsk and Kostroma (Fig. 8 a,b), and in moderately moist climates as in Tulun and Ogurtsovo (Fig. 8 e,f). For dry stations like Uralsk and

Yershov (Fig. 8 c,d), the described algorithm of frozen soil physics seem to work best. Interestingly, the scatter of model performance between different years for the same station may be significant. For example, for Uralsk (Fig. 8 c) the penetration of the freezing wave into deep layers of soil is generally slightly underestimated, but the freezing depth for the second year is overestimated. Such behavior can be found in the results from other stations. This can be explained by use of an overly simplified treatment of thermodynamical processes in soil, and the shortcomings of the empirical formula of the characteristic freezing curve (Eq. (9)). This relationship is, in fact, a much more complicated function, depending not only on soil properties, but on many other parameters such as the solute concentration in the soil solutions. The correlation of freezing depth simulation with the amount of soil water suggests that soil moisture might be another important factor in the definition of liquid mass transformation rate (Eq. (6)).

4. Performance of the coupled atmospheric/soil version of MAPS

In April 1996, the multi-level soil/vegetation model was introduced into the continuously running MAPS assimilation system. The soil temperature and volumetric water content fields, as predicted by the soil model, were allowed to evolve in the MAPS 3-hourly assimilation cycle for 1 year to May 1997. At that point, they began to evolve in a 1-hour assimilation cycle over the 6 months up to the beginning of the 9-month period (November 1997) for which hydrological cycle budget components in the Mississippi River basin were studied. The assimilation frequency was also 1 hour over the 9-month study period. Because there is not yet a high-frequency, national domain precipitation analysis available in real time, it is necessary to depend on the MAPS hourly precipitation forecasts for precipitation input.

Since January 1997, a snow model with accumulation and melting processes and a full energy budget has been running in the real-time MAPS. This scheme was made possible by the addition in the same month of a relatively sophisticated cloud microphysics scheme [the level 4 scheme from the NCAR/Penn State MM5 research model, *Reisner et al.*, 1998, *Brown et al.*, 1998], allowing for the formation, transport and fallout of cloud water and cloud ice as well as rain, snow, graupel, and the number concentration of cloud ice particles. The scheme assumes an exponential distribution in size of precipitation particles and permits the coexistence of both water and ice hydrometeors at a grid point, if the temperature is between 0° and -40°C. Along with the introduction of this scheme, a hydrometeor cycling capability has been added to MAPS, so that cloud fields from the previous 1-hour forecast are used to initialize each new forecast, minimizing cloud spin-up.

From January through March 1997, the snow fields in MAPS were allowed to cycle over each 24-hour period, with an update of the snow depth field occurring once daily from the US Air Force (USAF) snow cover analysis. That analysis was a large improvement over using no snow cover at all, but has problems in certain situations such as continuous low cloud cover. From early March 1997 through the end of May, the USAF analyses were unavailable and, consequently, the snow cover field in MAPS was allowed to cycle independently, driven by predicted snow accumulation and melting, just as was done with the soil moisture and temperature fields. The results of this ‘test’ (forced by external circumstances) were very satisfactory, and led us to allow snow water equivalent depth and snow temperature to continue to evolve based solely on MAPS forecasts. We suggest that even with an improved snow analysis in the future, model forecast snow information should be combined with observation-based analyses to determine optimal snow fields.

In April 1998 (during the 9-month study period), the frozen soil physics package described in section 5 was incorporated into the real-time MAPS forecast cycle. For the remainder of the spring, the effect of this change was to retard the warming in the northern part of the MAPS domain, where soil temperatures were still below freezing, and to increase runoff where snow cover was still present. The effects of frozen soil physics should be more significant in its first full winter season, the winter of 1998-1999.

4.1 Comparisons of monthly values between MAPS grids and external data

First, we examine the climatology of MAPS precipitation forecasts compared to monthly precipitation analyses from the National Climatic Data Center (NCDC). For brevity, comparisons only for the months of November 1997, February 1998, and April 1998 are shown in Fig. 9. The MAPS fields are summations of 6-9 hour forecasts for each 3-hour period over the entire month. For November 1998 (Fig. 9a), there was a general agreement of the spatial patterns with maxima on the West Coast of the United States and in the southern and eastern US. The observed (NCDC) patterns in orographic precipitation along the Cascades in the Pacific Northwest states were reflected well in the MAPS. Sometimes, finer details even showed good agreement, such as an east-west axis of 2-3 inches of precipitation along the Kansas-Nebraska border. The axis of heavier precipitation from West Virginia into Pennsylvania along the Appalachians appear both in the MAPS forecast field and in the NCDC analysis. However, there were also some consistent errors in the MAPS fields, such as along the Gulf Coast westward into northern Louisiana, where MAPS showed less than 2 inches of precipitation and over 7 inches was observed.

For February 1998 (Fig. 9b), the NCDC analysis showed heavier precipitation along the West Coast and in the southeast US than in November. The upper midwestern states remained fairly dry, consistent with the El Nino - Southern Oscillation (ENSO) event dominating the US winter precipitation patterns. These patterns were well-reflected in the MAPS monthly forecast, except for a general underforecast again in the southeast US. Some local anomalies in the observed patterns such as in North Dakota, Utah, and along the Mississippi River bordering Illinois and Missouri are also captured in the MAPS forecast. By April 1998 (Fig. 9c), precipitation had ended on the West Coast and in Florida and was heaviest in the southern Appalachians, patterns also shown in the MAPS forecasts. For this month, the MAPS forecast in the southeast US showed somewhat better agreement than in previous months.

While these comparisons show MAPS forecasts for the 6-9 h period, the evolution of soil moisture in the hourly MAPS assimilation cycle is actually controlled by the 0-1 hour precipitation forecast. Thus, it is important to examine the climatology of MAPS precipitation forecasts for different forecast durations. For a 12-day period during summer 1998, MAPS forecast precipitation was calculated for the periods of 0-1 hour, 0-3 hours, 3-6 hours, 6-9 hours, and 9-12 hours (Fig. 10). The variations in these different summations for this period are an indicator of the susceptibility of the MAPS forecast model to precipitation spin-up, a common problem in atmospheric prediction models. The same spatial patterns are certainly evident in all 5 summations, but some degree of increase in total forecast precipitation for the 12-day period is evident between the 0-1 hour period and periods further from the model initial time. The underforecast of the 0-1 hour period appears to range from 50% (e.g., southern Appalachians) to very small (e.g., northeastern Colorado). Overall, this underforecast will of course have a significant impact on the evolving MAPS soil moisture fields. It may be expected that spatial patterns of soil moisture from MAPS will show good reliability except in regions of systematic error such as near the Gulf Coast during winter. However, in our experience, this degree of spin-up is relatively small compared to other operational numerical models. Improvement in data assimilation of moisture- and cloud-related observations is probably the most important factor in decreasing these biases, although modifications in model physics are also needed. The needed modifications suggested by this result are discussed further in section 5.

The top layer (2.5 cm thick) and total (0-3 m) soil moisture from MAPS averaged for the month of April 1998 are presented in Fig. 11. Dry areas in Florida, Minnesota through Michigan's Upper Peninsula, the High Plains from Colorado/Kansas through Montana, and much of the southwest US are shown in both the top layer soil moisture and observed precipitation (NCDC) fields. The total soil moisture is much less responsive, as expected, to recent precipitation history. Observations of soil moisture over large regions are generally unavailable at the current time. The Palmer Drought Severity Index provides an indication of soil moisture, but only relative to climatology whereas the other parts of Fig. 11 are absolute measures. The Palmer Index for April 1998 shows some agreement for drought regions from North Dakota westward into Montana with the absolute measures, but is difficult to correlate otherwise.

A comparison of monthly surface temperatures is also made (Fig. 12) between NCDC analyses and MAPS forecasts. Here, the agreement is quite good, except for some regions of the western US where the difference is attributable to elevation differences between stations used in NCDC analyses and the actual mean elevation of the areas (closer to temperatures consistent with MAPS fields).

4.2 Seasonal variations of components of the hydrological cycle from MAPS

Several different components of the hydrological cycle were examined for mutual consistency over the entire MAPS domain, including precipitation minus evaporation (P-E), snow accumulation, surface runoff, and snow melt. These fields are shown in Fig. 13 for the months of February and May 1998. In February 1998, the precipitation-evaporation difference is positive throughout most of the MAPS domain, and particularly so along the West Coast, in the lower Mississippi Valley, and along the East Coast. By May, this field had become negative, meaning a tendency to soil drying, over the majority of the MAPS domain. The evolution of the other fields for both months appear to be physically consistent and reasonable. For example, largest amounts of runoff appear where P-E is also large (although the temporal distribution of the precipitation is also clearly important, with concentration in a few intense episodes more likely to produce high runoff amounts). The snow melt field in February extends fairly far south over higher terrain regions in the western and eastern US, but is non-zero only in the highest terrain regions in the West by May. On average snow is usually found within the MAPS domain in May only over the highest elevations of the western mountains.

Finally, we present area-averaged hydrological cycle components from MAPS for the entire November 1997 - July 1998 period for four quadrants occupying much of the Mississippi River basin (Fig. 14a). These quadrants are chosen somewhat arbitrarily, but one may expect certain climatological features to be reflected in the MAPS fields, such as more snow in the northern quadrants, and drier conditions in the western quadrants. Other areas could be studied, as needed, from the MAPS MORDS data sets for GCIP.

The winter part of this period was dominated by a strong ENSO event, with much warmer than normal temperatures across the northern two thirds of the US [Climate Prediction Center and NCDC analyses not shown here]. The precipitation from the same analyses showed somewhat drier than normal conditions in the same area, which covers quadrants A and B and over half of quadrants C and D.

The evolution of these area-averaged hydrological cycle components (Fig. 14 b-e) indicate a general increase of precipitation toward summer, and maximum drying (minimum P-E) in May.

The P-E term goes from positive to negative in all 4 quadrants between March and April, one month earlier than the 20-year mean value over much of the US calculated by Ropelewski and Yarosh (1997). Precipitation for this 9-month period is clearly lowest overall in the northwest quadrant (A - northern high plains). The southeastern quadrant (B - central Mississippi Valley) has the most precipitation overall, and peaks in runoff in March and July. Snow melt peaks in the two western quadrants in March, and appears to contribute to a peak in runoff in the southwest quadrant (containing the Colorado Rockies). In quadrant A (northwest), most of the precipitation in February and March is snow, but consistent with the ENSO event, the majority of the winter precipitation in quadrant B (northeast - Minnesota, Wisconsin) was liquid, a very anomalous year in this regard. The precipitation was quite low through May (when the ENSO event was ending) in 3 of the 4 quadrants, excepting quadrant D (Missouri, Illinois, central Mississippi Valley). Only in quadrant D was there significant precipitation throughout the winter. Surface runoff was also much larger in quadrant D than the other areas. The surface runoff would likely have been larger in quadrants A and B if the frozen soil physics had been in place before April.

Overall, analyses like those shown in Fig. 14 allow evaluations of these hydrological cycle components in a time-continuous and physically consistent manner (within the limitations of the coupled model and associated data assimilation). Multi-year records of such fields from MAPS and other models are now being created as part of GCIP, and multi-year studies from these fields are likely to provide improved understanding of longer-term variations in the hydrological cycle.

5. Concluding remarks

This study documents the current progress and relative success in using a mesoscale atmospheric/land-surface coupled model with high-frequency assimilation of atmospheric observations

(MAPS) to produce physically consistent fields of soil variables and hydrological cycle components. The assimilation frequency in MAPS is quite high (1 hour since May 1997) and higher than other data assimilation systems currently used in operational meteorological forecast centers. This constrains atmospheric model drift in regions where data are plentiful. The assimilation approach in MAPS for soil and snow fields is to produce them through continuous cycling rather than to impose them from external observations. This allows these fields to show good internal time consistency as well as consistency with atmospheric fields. The success of this approach depends critically on the use of a land-surface parameterization with minimal internal drift or bias.

The hydrological budget of the Mississippi River basin, as much of the global land area, is dominated much of the year by processes related to snow and sub-freezing temperatures in the soil. To improve the handling of these processes in the MAPS coupled model, parameterizations for snow and frozen soil have been added and described in detail in this paper. One-dimensional tests of this version of the MAPS land-surface model have been performed using observation data sets of up to 18 years from six sites in Russia and one in Kazakhstan. One of these sites, Valdai, Russia, is the focus of the PILPS 2d test. Overall, the MAPS 1-D model gave good performance for these sites in forecasts of soil temperature, soil moisture, and freezing level. The main features in seasonal change of soil moisture and in snow accumulation and melting were captured well. The soil freezing level simulations from MAPS showed some errors due to insufficient vertical resolution and simplicity in the characteristic freezing curve, including neglect of potential factors such as solute concentration. Tests without the treatment of frozen soil physics showed that this addition to the MAPS land-surface process model has given positive results in simulation of hydrological cycle components. Further improvement may be achieved by more accurate treatment of soil properties in the frozen and unfrozen soil with the possibility of changing these properties with depth, and also by defining snow characteristics as a function of the snow age.

The multi-level soil/vegetation model in MAPS has been cycling soil fields since April 1996 and snow fields since March 1997. Comparisons were made between monthly fields of mean temperature and accumulated precipitation from MAPS and NCDC analyses for a 9-month period from November 1997 through July 1998, showing very good agreement for temperature and fairly good agreement for precipitation. Many details in precipitation fields were captured by MAPS, including those related to orographic precipitation. One systematic problem was also shown in that too little precipitation winter precipitation near the Gulf Coast was forecast by MAPS. This problem appears to be related to overforecasting of convective precipitation over warm water, indicating a correction required in the feedback between ocean surface fluxes and the convection parameterization in MAPS. A comparison of MAPS forecast precipitation from different time projections over a 12-day period also revealed some improvement needed in the precipitation spin-up, but this is not a crippling problem. Ongoing work to assimilate satellite, radar, and surface data into MAPS cloud/moisture analyses is expected to alleviate this problem somewhat. Qualitative verification of soil moisture, surface runoff, precipitation-minus evaporation, snow accumulation, and snow melt fields over the 9-month test period shows that these fields, in general, are quite realistic, with good time continuity and mutual consistency. Intercomparison of analyses between different models including MAPS (*Berbery et al*., 1999) shows generally good results for MAPS but also indicates possible need for improvement in surface physics, cloud description, and radiation.

The key areas of focus, many indicated in this study, in MAPS development over the next two years are assimilation of cloud/precipitation observations, further improvements to atmospheric surface layer and soil physics, use of improved soil/vegetation data sets available that cover the MAPS domain, and improvements to the MAPS convective precipitation parameterization.

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Appendix

List of main symbols

a, b, c Empirical dimensionless factors dependent on soil type

c_p Specific heat capacity of air under constant pressure $[J kg^{-1} K^{-1}]$

c_{sn} Specific heat capacity of snow $[J kg^{-1} K^{-1}]$

C Volumetric heat capacity of the layer spanning the ground surface $[J m^{-3} K^{-1}]$

C_a Apparent volumetric heat capacity of soil $[J m^{-3} K^{-1}]$

C_i Volumetric heat capacity of ice $[J m^{-3} K^{-1}]$

C_l Volumetric heat capacity of liquid water $[J m^{-3} K^{-1}]$

C_s Volumetric heat capacity of dry soil $[J m^{-3} K^{-1}]$

C_{sn} Volumetric heat capacity of soil $[J m^{-3} K^{-1}]$

C^* Canopy water content $[m]$

D Water drip rate from canopy to soil $[kg\ m^{-2} s^{-1}]$

D_f Diffusional conductivity for frozen soil $[m^2 s^{-1}]$

E Surface flux of total moisture content in the atmosphere $[kg\ m^{-2} s^{-1}]$

E_c Evaporation flux from the canopy $[kg\ m^{-2} s^{-1}]$

E_{dir} Evaporation flux from the bare soil $[kg\ m^{-2} s^{-1}]$

E_t Transpiration flux $[kg\ m^{-2} s^{-1}]$

F Heat of snow melting $[W\ m^{-2}]$

g Acceleration of gravity $[m\ s^{-2}]$

G_{sn} Heat flux into the snow $[W\ m^{-2}]$

H Sensible heat flux from ground $[W\ m^{-2}]$

H_{rain} Heat brought to the ground surface by liquid phase of precipitation $[W\ m^{-2}]$

h_{sn} Snow depth $[m]$

I_m Infiltration flux from snowmelt $[kg\ m^{-2} s^{-1}]$

K_f Hydraulic conductivity in frozen soil $[m\,s^{-1}]$

K_s Saturated soil value of hydraulic conductivity $[m\,s^{-1}]$

L_f Latent heat of fusion $J kg^{-1}$

L_s Latent heat of sublimation $J kg^{-1}$

P_l Flux of liquid precipitation $[kg\ m^{-2} s^{-1}]$

R_n Net radiation $[W\ m^{-2}]$

S_{li} Rate of liquid mass transformation into ice $[kg\ m^{-3} s^{-1}]$

S' Saturation water content for a canopy surface (=0.005 m)

T Temperature $[K]$

T_{sn} Temperature at the snow surface $[K]$

T_s Temperature at the soil-snow interface $[K]$

Θ Density of total soil moisture content $[kg\ m^{-3}]$

Θ_r Density of minimum total soil moisture content $[kg\ m^{-3}]$

Θ_s Density of maximum total soil moisture content $[kg\ m^{-3}]$

W_s Moisture flux into the ground $[kg\ m^{-2} s^{-1}]$

z	Vertical coordinate, increasing upward $[m]$
Δz_a	Half of the lowest model level height (5 m)
Δz_{sn}	Half of the snow depth $[m]$
Δz_s	Half of the top soil layer depth $[m]$
η	Volumetric water content of soil (dimensionless).
η_g	Volumetric water content of soil at the ground surface
η_i	Volumetric content of ice in soil
η_l	Volumetric content of liquid phase in soil
η_s	Porosity of soil
ν	Thermal conductivity of soil $[W\ m^{-1} K^{-1}]$
ν_f	Thermal conductivity of potentially frozen soil $[W\ m^{-1} K^{-1}]$
ν_{sn}	Thermal conductivity of snow $[W\ m^{-1} K^{-1}]$
Ψ_s	Moisture potential for saturated soil $[m]$
Ψ_f	Moisture potential for partially frozen soil $[m]$
ρ_a	Air density at the lowest model level $[kg\ m^{-3}]$
ρ_i	Density of ice $[kg\ m^{-3}]$
ρ_l	Density of liquid water $[kg\ m^{-3}]$
ρ_s	Soil density $[kg\ m^{-3}]$
ρ_{sn}	Snow density $[kg\ m^{-3}]$
σ_f	Non-dimensional plant shading factor

Figure Captions

Figure 1. A summary of the processes in the MAPS/RUC soil/snow/vegetation scheme.

Figure 2. Soil temperature profiles simulated by MAPS with parameterization of frozen soil physics (solid line), and without parameterization of frozen soil physics (dot-dashed line). 15 April 1981, Valdai, Russia.

Figure 3. Profiles simulated by MAPS 1-D land-surface model with parameterization of frozen soil physics (solid line) and without (dot-dashed line) for (a) Apparent heat capacity, and (b) thermal diffusivity. 15 April 1981, Valdai, Russia.

Figure 4. Volumetric soil moisture content profiles simulated by MAPS with parameterization of frozen soil physics (solid line), and without (dot-dashed line) for (a) 15 January, (b) 15 April, (c) 15 July, and (d) 15 October for 1981, Valdai, Russia.

Figure 5. Profiles simulated by MAPS 1-D land-surface model with parameterization of frozen soil physics (solid line) and without (dot-dashed line) for (a) Diffusional conductivity and (b) hydraulic conductivity. 15 April 1981, Valdai, Russia.

Figure 6. Annual variation of variables averaged in MAPS simulations over 18-year period and observations for Valdai, Russia. (a) Skin temperature, (b) snow water equivalent, (c) monthly accumulated total evaporation, (d) soil moisture content in the top 1-m layer, (e) total and surface runoff

Figure 7. Observed and MAPS simulated soil moisture (mm) in top 1-m layer of soil (1978-1983) for six stations for (a) Khabarovsk, Russia, (b) Kostroma, Russia, (c) Uralsk, Kazakhstan, (d) Yershov, Russia, (e) Tulun, Russia, (f) Ogurtsovo, Russia

Figure 8. Observations and MAPS simulations of freezing depth (cm) in soil for same stations and period as in Figure 7.

Figure 9. Monthly accumulated precipitation from NCDC analyses and series of MAPS 6-9 hour forecasts for November 1997, February 1998, and April 1998.

Figure 10. Accumulated precipitation from MAPS model for different projection periods. Period is for 23 July - 3 August 1998.

Figure 11. Mean soil moisture from MAPS for April 1998 compared with NCDC precipitation analysis and Palmer Drought Severity Index for same month.

Figure 12. Mean surface temperature from NCDC analyses and MAPS analyses for February 1998 and April 1998.

Figure 13. Hydrological cycle components from MAPS 6-9 h forecasts for (a) February 1998 and (b) April 1998. Shown are precipitation minus evaporation, snow accumulation, surface run-off, and snow melt.

Figure 14. Monthly values (November 1997 - July 1998) of area-averaged hydrological cycle components from MAPS for four geographical areas. (a) map of four areas (A, B, C, D), (b)-(e) are for areas A, B, C, D.